# Chapter 1

# Introduction.

The large-scale geodynamics of the present-day Earth is to a large extent connected to plate tectonics. The formation of oceanic lithosphere at the spreading centers of mid-ocean ridges and the subduction of this lithosphere at subduction zones are the major expressions of this mechanism. It is ultimately driven by a temperature difference between the surface of the Earth and its interior, and efficiently transports heat to the surface, where it is transferred to the hydrosphere and atmosphere and finally radiated into space. This effectively cools the Earth's interior.

At mid-ocean ridges, new basaltic crust is formed by solidification of melt which is produced by pressure release partial melting in the hot upwelling mantle below the ridge. Much heat is released at this point through conduction and hydrothermal circulation (Stein, 1995), and the newly formed crust continues to cool as it moves away from the ridge to make place for new basaltic crust. This makes the plates an important factor in the cooling of the planet, facilitating about 70 % of the global surface heat flux (Pollack et al., 1993). The main force which drives this mechanism is *slab pull*, i.e. the force exerted by a negatively buoyant slab that is subducting (Stacey, 1992). When oceanic crust is created at mid ocean ridges, its intrinsic lower density (of both basalt and harzburgite relative to fertile peridotite) and relatively high temperature cause it to be positively buoyant. As it cools on its way from the ridge to a subduction zone, density increases, and at some point, the buoyancy becomes negative. This negative buoyancy is in fact the cause of the slab pull. A secondary driving force is ridge push, which is an order of magnitude weaker than slab pull (Turcotte and Schubert, 2002).

At subduction zones, the oceanic lithosphere sinks into the mantle below another lithosphere. As it moves downward, it heats up and fluids are released from the crust, which was hydrated by hydrothermal circulation of ocean water. This facilitates partial melting in the overlying mantle wedge, leading to the formation of a volcanic arc parallel to the subduction zone. This process itself at a continental margin or the accretion of these arcs in an oceanic case to existing continental masses is seen as an important mechanism for continental growth in the present-day Earth, together with the accretion of oceanic plateaux and intraplate volcanism (Rudnick, 1995).

Early in the planet's 4.57 billion year history, its internal temperature was higher by some hundreds of degrees Celsius than at present. This heat originated mainly from three sources. The first is the release of potential (gravitational) energy during accretion of the Earth. This took place over a period of 50 to 100 million years (Allègre et al., 1995; Halliday and Lee, 1999), providing enough energy to completely melt the planet (Horedt, 1980). The second heat source is the release of gravitational energy by core differentiation. The mean age of core differentiation was determined by Pb-Pb isotope geochemistry at 4.45 Ga (Allègre et al., 1995). The amount of heat released was sufficient to raise the core temperature by some thousands of degrees (Horedt, 1980). The decay of radioactive elements forms the third heat source, which is the most important for long-term development (Stevenson, 2003). Short-lived isotopes, specifically $^{244}$Pu, $^{236}$U and $^{26}$Al with half lives of 76 Myr, 24 Myr and 0.72 Myr, respectively, and longer-lived isotopes such as $^{40}$K, $^{235}$U, $^{238}$U, and $^{232}$Th added large amounts of heat. $^{26}$Al by itself produced enough energy to raise the core temperature by 10000 K (Ruff and Anderson, 1980). The decay of radioactive isotopes continues to provide heat at an ever decreasing rate. This elevated internal temperature had a major impact on the process of the formation of lithosphere from upwelling mantle flows as it occurs today at mid-ocean ridges. At a higher mantle temperature, upwelling material crosses the solidus at greater depth and therefore has a longer melting path, producing a thicker basaltic crust and underlying Harzburgite layer (Vlaar, 1986; Vlaar and Van den Berg, 1991). Clearly, a thicker crust can only be produced when upwelling and melting can take place over the complete depth range between solidus and surface, as is approximately the case at mid ocean ridges presently. In situations where upwelling material cannot rise to shallow levels, the production of greater crustal thicknesses is mitigated.

Since both basalt and harzburgite have a lower density than the undepleted peridotite mantle source rock, the layering that is created at mid-ocean ridges is gravitationally intrinsically stable. Cooling of this package, however, increases its density up to a point where the lithosphere is no longer positively buoyant, after which active subduction can take place. For the present-day Earth, this is at about 20 to 40 million years (Oxburgh and Parmentier, 1977; Vlaar and Van den Berg, 1991; Davies, 1992) after formation. For a thicker pile which is produced in a hotter mantle (Sleep and Windley, 1982; Vlaar, 1986; Vlaar and Van den Berg, 1991), this neutral buoyancy age is increased up to several hundreds of million years or more (Vlaar and Van den Berg, 1991). This hinders plate tectonics as we observe it in the present-day Earth. Numerical models of a subduction system by Van Hunen (2001) show that the present-day subduction style is feasible in a mantle that is up to 150 Kelvin hotter than it is today. At still higher mantle temperatures, these models show that the slab breaks off at very shallow levels, due to a combined effect of weaker rheology and increased buoyancy of a thicker partly eclogitic crust, removing slab pull and causing a rebound.

Our closest neighbours in the solar system, Mars and Venus are comparable in size and composition to Earth, and may be or have been subject to the same processes. Although both planets do not have plate tectonics today (Zuber, 2001; Nimmo and McKenzie, 1996), it may have been active during their earlier histories. Sleep (1994) interprets the current geology of Mars in a framework of early plate tectonics on this planet. Satel-

| potential temperature ($^oC$) | viscosity | scale factor |
|---|---|---|
| 1350 | 1.000 | 1.000 |
| 1450 | 0.258 | 0.713 |
| 1550 | 0.077 | 0.523 |
| 1650 | 0.026 | 0.402 |
| 1750 | 0.010 | 0.316 |

Table 1.1: The scaling of the magnitude and length scale of temperature variations in the mantle as a function of potential mantle temperature relative to the present temperature of about 1350$^o$C, using the scaling relations of McKenzie and Bickle (1988). The viscosity is computed assuming diffusion creep only and using an activation energy of 300 kJmol$^{-1}$ and an activation volume of $6 \cdot 10^{-6}$ m$^3$mol$^{-1}$ (Karato and Wu, 1993) at a depth of 300 km.

lite measurements of crustal magnetization on the southern hemisphere of Mars show a pattern more or less similar to magnetic striping found on the Earth's ocean floor. Although both the width and signal magnitude are much larger than on Earth, a similar plate tectonic origin has been suggested (Acuña et al., 1999; Connerney et al., 1999). Crater counts indicate that Venus has undergone a global resurfacing between 300 and 600 Myr before present (Schaber et al., 1992; Nimmo and McKenzie, 1998). We therefore do not have evidence for processes prior to the resurfacing, but episodic plate tectonics has been suggested for this planet (Turcotte, 1993).

Because viscosity is strongly temperature dependent, a higher mantle temperature in the early Earth (or other planets) also has consequences for the dynamics of mantle convection. A reduced viscosity results in faster flow and reduces the buildup of large temperature contrasts. McKenzie and Bickle (1988) use the boundary layer analysis of Turcotte and Oxburgh (1967) to estimate that the magnitude and length scale of temperature variations scale with the viscosity to the power $\frac{1}{4}$. From this we find that for a mantle that is 100 K hotter than it is today, both the scale and the magnitude of the variations would be reduced to about 70 percent of the present values. For still higher mantle temperatures, this reduces further down to about 30 percent for a 400 K hotter mantle, see Table 1.1 (a similar argument with an even larger decrease in viscosity and therefore variations is given in Nisbet et al., 1993). In other words, this line of reasoning predicts that in a hotter mantle circulation patterns have a smaller scale than the $x \cdot 10^3$ km of ridge to subduction zone which is the surface expression of present-day mantle convection. Furthermore, hypothetical mantle plumes, associated with hotspots like Hawaii and Iceland, with excess temperatures of up to 200 to $250^oC$ (Herzberg and O'Hara, 1998), would have both a smaller size and a reduced excess temperature in a hotter mantle. Direct evidence for hotter conditions in the early Earth comes from Archean geology (e.g. Nisbet et al., 1993).

In several parts of the world, rocks which have been formed during the Archean (3.8-2.5 Gyr) outcrop at the surface. An overview of these locations is shown in Figure 1.1.

Figure 1.1: Overview of Archean cratons on the Earth. Black indicates outcropping Archean rocks. After Condie (1981).

Evident differences in the geology of these cratons compared to continental areas of more recent age indicate different conditions of formation. Goodwin (1991) gives an overview of the characteristics of (preserved) Archean crust, which will be summarized here. The cratons are typically subcircular to oblong and have unexposed extensions. They generally consist of about 60% granitoid gneiss-migmatites, including substantial granulites. The presence of high grade metamorphic xenoliths originating from upper crustal material in these rocks indicates still older crust. Geochemistry of the gneisses points to a close mantle-derived petrogenesis (low Sr and Nd initial ratios), and a relatively deep melt origin of more than 40 km (high La/Yb ratios), except for the latest Archean igneous rocks, which are formed by melting in the crust itself (Eu depletion). 30% of the cratons is made up by massive granitoid plutons. The older plutonic rocks generally classify as tonalite-trondhjemite-granodiorite (TTG) suite, and the younger are characteristically potassic granites. Also, small but significant amounts of layered mafic to ultramafic intrusives are found. Greenstone belts form about 10% of the exposed Archean rocks. These generally consist of tholeiitic to komatiitic lava flow sequences on top of which are felsic pyroclastic rocks (the former and the latter in varying proportions) and sediments. Most greenstone belts have been formed either around 3.5 Ga or around 2.7 Ga.

The style of tectonic deformation during the Archean shows a large horizontal component, with later vertical motions. Also, the intrusion of diapirs forming dome structures is important. An example of the resulting structures can be seen in Figure 1.2, which shows

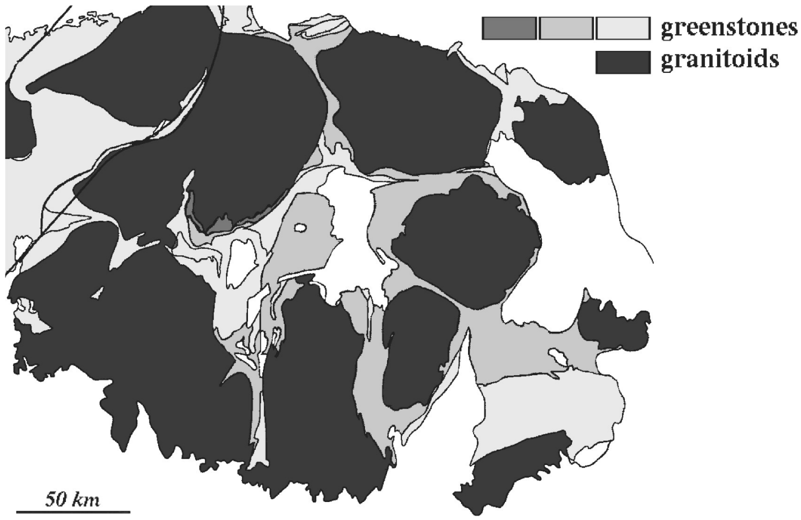


Figure 1.2: Overview map of the eastern Pilbara terrane, showing a typical structure of granites surrounded by greenstone belts (modified from Zegers and Van Keken, 2001).

a simplified geologic map of the Archean eastern Pilbara region in Australia, consisting largely of granite bodies of 50 to 100 km diameter, surrounded by greenstones. Seismological determination of upper mantle structure tells us that Archean cratons are usually underlain by thick roots down to about 250 km depth or deeper (Grand, 1994; Trampert and Woodhouse, 1996; Fischer and Van der Hilst, 1999), characterized by high seismic velocities. They are thought to consist of relatively cold and depleted mantle peridotite, and are often invoked to explain the long-term stability of these cratons (Forte and Perry, 2000; De Smet et al., 2000a). Numerical modelling experiments have shown that mantle diapirs could be important agents in the growth of continental roots during the Archean and Proterozoic (De Smet et al., 1998; Van Thienen et al., 2003a, chapter 8). These mantle diapirs are generally on a scale of 50 to 100 km and penetrate an existing root while producing melt, and a complementary depleted low density residue, thus adding depleted material to the continental root.

These results from numerical models are confirmed by geological data. Samples of cratonic lithosphere are found as xenoliths in kimberlite intrusions (Nixon and Boyd, 1973) and as larger bodies in some orogenic belts (Brueckner and Medaris, 1998; Van Roermund and Drury, 1998). It is proposed that many of these mantle rocks are emplaced into the lithosphere by diapirs from the convecting sublithospheric mantle (Nicolas, 1986) in a variety of geodynamic environments (Green and Gueguen, 1983; Nicolas et al., 1987; Fabriès et al., 1991). The PT paths derived from some cratonic peridotites from the Norwegian Western Gneiss Region (Van Roermund and Drury, 1998; Drury et al., 2001) imply that these rocks were part of diapirs that intruded cratonic lithosphere of Archean to early Proterozoic age (the cooling age of these rocks is 1.7-1.8 Ga, but the Sm-Nd model age indicates a depletion age of 2.5-3.0 Ga). Drury et al. (2001) have shown that the PT path of these peridotites is consistent with PT paths of diapiric upwellings calculated from the thermo-chemical convection models of De Smet et al. (2000a).

It is the aim of this thesis to investigate the types of geodynamics that may have been important under the strongly different thermal conditions of the early Earth, both in the cooling of the planet's interior and in the formation and recycling of oceanic crust and of the continents, and also in the evolution of Mars and Venus. Furthermore, the demise of these processes as the planets cool towards the present temperature are investigated.

The tools which are used in the research are numerical models. These include both relatively simple models based on parametric descriptions of processes and 2-D thermo-chemical mantle convection models including compositional differentiation by partial melting.

An overview of the contents of this thesis will be given below.

Following this introduction, the main physical concepts and model equations which describe the physics are presented in chapter 2. The numerical techniques that are used to solve the model equations of chapter 2 are described in chapter 3. In following chapters, the numerical models presented in chapters 2 and 3 are applied to the problem area of this thesis. These chapters more or less show a progression in time and a zooming from large scale to smaller scale. First, the conditions under which different geodynamic regimes operate and the global thermal characteristics of these regimes are mapped out using simple models based on parameterizations of the relevant processes. In chapter 4, the effect of mantle temperature and gravitational acceleration on the time scales of the development of negative buoyancy of oceanic lithosphere is investigated in a plate tectonic setting for the terrestrial planets Earth, Venus and Mars. This gives an indication of the conditions that allow Earth-like plate tectonics to take place on the terrestrial planets. Next, in chapter 5, the cooling capacity and time characteristics of both plate tectonics and stagnant-lid basalt extrusion mechanism are assessed and compared, using a parametric approach. The following chapters consider in more detail the dynamics of crustal growth and recycling in a hot mantle. Numerical thermochemical mantle convection models on a scale of hundreds to 1500 km are used to investigate the local processes rather than the global effects. Chapter 6 deals with the formation of a basaltic oceanic crust from a convecting mantle and the recycling of the crust back into the mantle. Numerical thermochemical convection models are used in this chapter, and the focus is both on the dynamics of the processes and on the implications for mantle chemistry. In chapter 7, modelling results of self-consistent continental growth in a dynamical mantle convection model are presented. The dynamics and time and spatial scales of the process are investigated, and results are linked to the petrology and geochemistry of Archean continental material as found in the cratons. Chapter 8 deals with the effect of different rheological models on the development of small scale mantle diapirs impinging on a growing continental root. These plumes add relatively light material to the root of a continent as produced in chapter 7 and have a stabilizing effect on it. In the concluding chapter 9, a summary and general conclusions of this thesis will be given, from the interpretation of the combined results of the different chapters.